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Holocene peatland development along the eastern margin of the Tibetan Plateau

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ABSTRACT

Knowledge of peatland initiation, accumulation, and decline or cessation is critical in understanding peatland development and the related carbon source/sink effect. In this study, we investigated the development of three peat profiles along the eastern margin of the Tibetan Plateau (ETP) and compared the results with those of our previous work along this transect. Our work showed that the initiation over the northern ETP is later and the slowdown/cessation earlier than in the middle to southern ETP. The timing of optimum peatland formation over the northern ETP lags the Holocene climatic optimum. These spatio-temporal differences are likely to be related to the intensity of Asian summer monsoon. Our work suggests that some peatlands along the ETP transect have returned or are now returning their previously captured carbon to the atmosphere and thus act as carbon sources. Some peatlands still have net accumulation at present, but the rates have been reduced concomitant with the decreasing summer monsoon intensity. We speculate that more of the previously stored carbon in the ETP peatlands will be re-emitted to the atmosphere if the aridity continues, as might occur under a continuous global-warming scenario.

Introduction

Peatlands are important terrestrial ecosystems and carbon reservoirs. They are closely linked to the global/regional carbon cycles by sequestering CO_2 from and/or emitting CO_2 and CH_4 back to the atmosphere (Gorham, 1991). There is growing concern that peatlands will return the previously captured C to the atmosphere by releasing CO_2 and/or CH_4 in response to climatic change, which would possibly accelerate the present global warming. As a result, increasing attention has been paid to the development of peatland and the related carbon budgets during the past decade (e.g. Smith et al., 2004; MacDonald et al., 2006; Jones and Yu. 2010; Flanagan and Syed, 2011; Reyes and Cooke, 2011). If peatlands are significant and fragile carbon reservoirs, it is important to study them in as wide a geographic context as possible, in order to help improve the accuracy of the estimated global carbon budget.

The eastern margin of the Tibetan Plateau (ETP) is influenced by the East Asian summer monsoon, the Indian summer monsoon, the East Asian winter monsoon, and the westerly jet stream. The modern climate generally ranges from humid/semi-humid to arid/semi-arid from south to north along the ETP (Fig. 1). Previous work has suggested that precipitation is an essential determinant of plant growth on the ETP (Hong et al., 2003; Xu et al., 2006, 2007; Zhou et al., 2010; Zhao et al., 2011). Long-term precipitation from the Asian summer monsoon in the ETP is thought to be closely related to the orbital solar insolation (e.g. An et al., 2011), while Holocene climate there is subject to millennial/centennial oscillations (Hong et al., 2003; Dykoski et al., 2005). However, the timing and amplitudes of the climatic fluctuations are inconsistent from region to region of the ETP (Hong et al., 2003; Wang et al., 2005; Dykoski et al., 2005; Shen et al., 2005; Zhou et al., 2010; Zhao et al., 2011). For example, the Holocene thermal maximum (HTM), widely recognized as a warm and humid period in monsoonal Asia, exhibits spatial and temporal asymmetry between different regions (Shi et al., 1992; An et al. 2000; Renssen et al., 2009; Zhao et al., 2009). These climatic asymmetries would result in different initiation timing and accumulation rate of peatlands, and therefore lead to different carbon budgets between different regions. Study on the control of peatland development along the ETP transect is important to shed light on the responses/feedbacks of peat growth to global climate changes and the related carbon source/sink effect.

In this study, we investigated the development of three peat profiles, namely the Ebo and Laji profiles in northern ETP and the Butuo peat core in southern ETP, and compared the results with those of our previous work at Hongyuan, the Zoige plateau, in the middle of the ETP (Fig. 1; Hong et al., 2003; Xu et al., 2006). The Ebo and Laji profiles are located in the Asian summer monsoon frontier area (arid/semiarid zone); the Butuo peat core is located in the typical Indian summer monsoon region (humid climate); and the Hongyuan profile is located in a transition area between the northern and southern ETP. These profiles serve as typical examples with which to constrain peatland development and the related carbon source/sink effects along the ETP transect.

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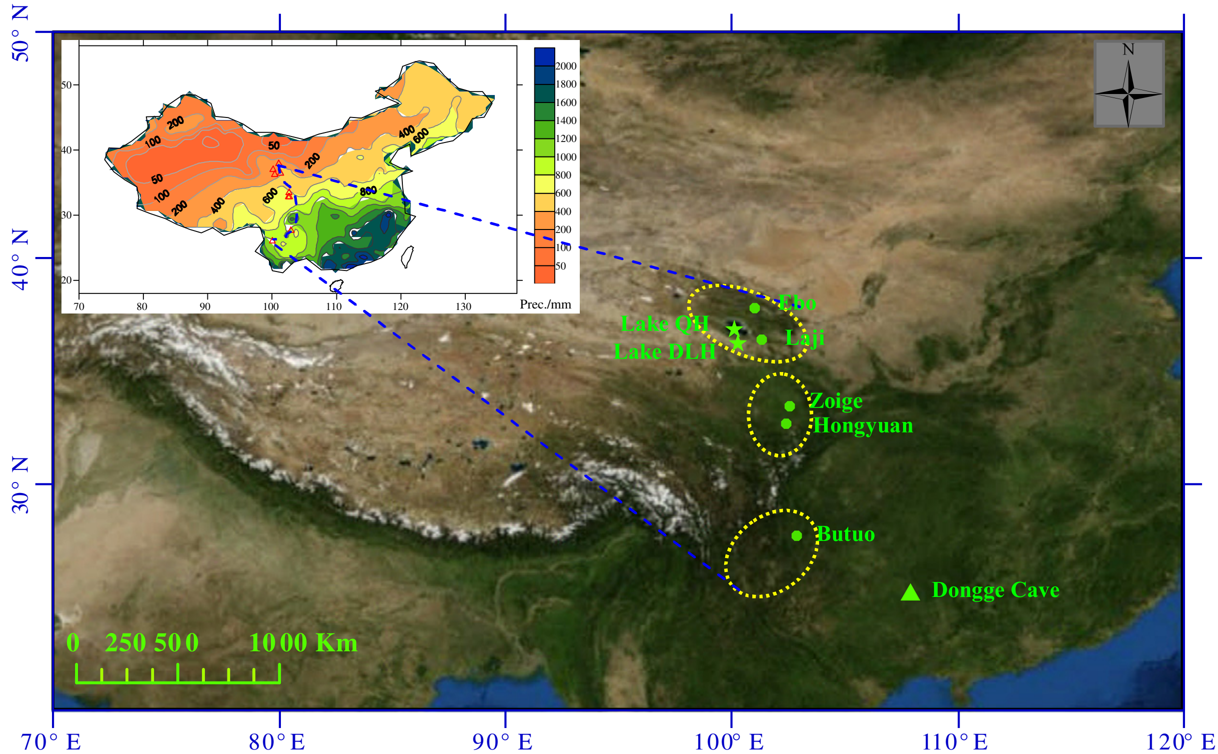

Figure 1. Location of the Ebo, Laji, and Butuo sampling sites. Also shown are sites mentioned in the text along the eastern margin of the Tibetan Plateau (ETP). The three yellow dotted ellipses symbolically indicate the climatic types along the ETP transect (arid/semiarid, transitional, and humid, from north to south). Map in upper corner shows the average annual precipitation ("Prec.") contours in China. Precipitation in the ETP is dominantly in the summer growing season.

Table 1
^{14}C and OSL ages of peatlands along the eastern margin of the Tibetan Plateau.

Profiles	Sample no.	Depth/cm	Dating materials	^{14}C ages ^{14}C yr BP	Error (1σ) (^{14}C yr)	Median calibrated ^{14}C ages[a] (cal yr BP)	OSL ages (ka)
Butuo	BT-43	43	TOM	2084	29	2055	
	BT-58	58	Cellulose	2441	28	2480	
	BT-80	80	Cellulose	3109	35	3338	
	BT-95	95	Cellulose	4210	35	4740	
	BT-107	107	Cellulose	5844	29	6665	
	BT-120	120	Cellulose	6201	38	7092	
	BT-150	150	Cellulose	6753	25	7608	
	BT-185	185	Cellulose	7652	33	8438	
	BT-196	196	Cellulose	8164	43	9104	
	BT-208	208	Cellulose	8612	32	9555	
	BT-230	230	Cellulose	10,717	52	12,630	
	BT-245	245	TOM	11,497	41	13,350	
	BT-258	258	TOM	11,747	212	13,603	
Laji	BT-285	285	TOM	14,544	48	17,708	
	LJ-3-51	51	Cellulose	6520	36	7439	
	LJ-3-80	80	Cellulose	6990	38	7828	
	LJ-3-100	100	Cellulose	6018	28	6860	
	LJOSL-1	20	Quartz				2.5 ± 0.1
	LJOSL-2	52	Quartz				2.7 ± 0.1
	LJOSL-3	75	Quartz				3.4 ± 0.1
	LJOSL-4	105	Quartz				5.1 ± 0.2
Ebo	EB-18	53–54	Cellulose	4634	23	5417	
	EB-26	*69–70*	*Cellulose*	*4650*	*40*	*5404*	
	EB-36	*89–90*	*Cellulose*	*6450*	*40*	*7367*	
	EB-37	91–92	Cellulose	6601	30	7494	
	EBOSL-1	30	Quartz				1.5 ± 0.1
	EBOSL-2	130	Quartz				117.1 ± 5

The italic underlined data (*EB-26*, *EB-36*) were measured at the Beta Analytic Radiocarbon Dating Laboratory, while the others were measured at the Institute of Earth Environment, Chinese Academy of Sciences (IEECAS).
[a] Calibrated ^{14}C ages are median probabilities (1σ) calibrated by the software of *Calib 6.02*.

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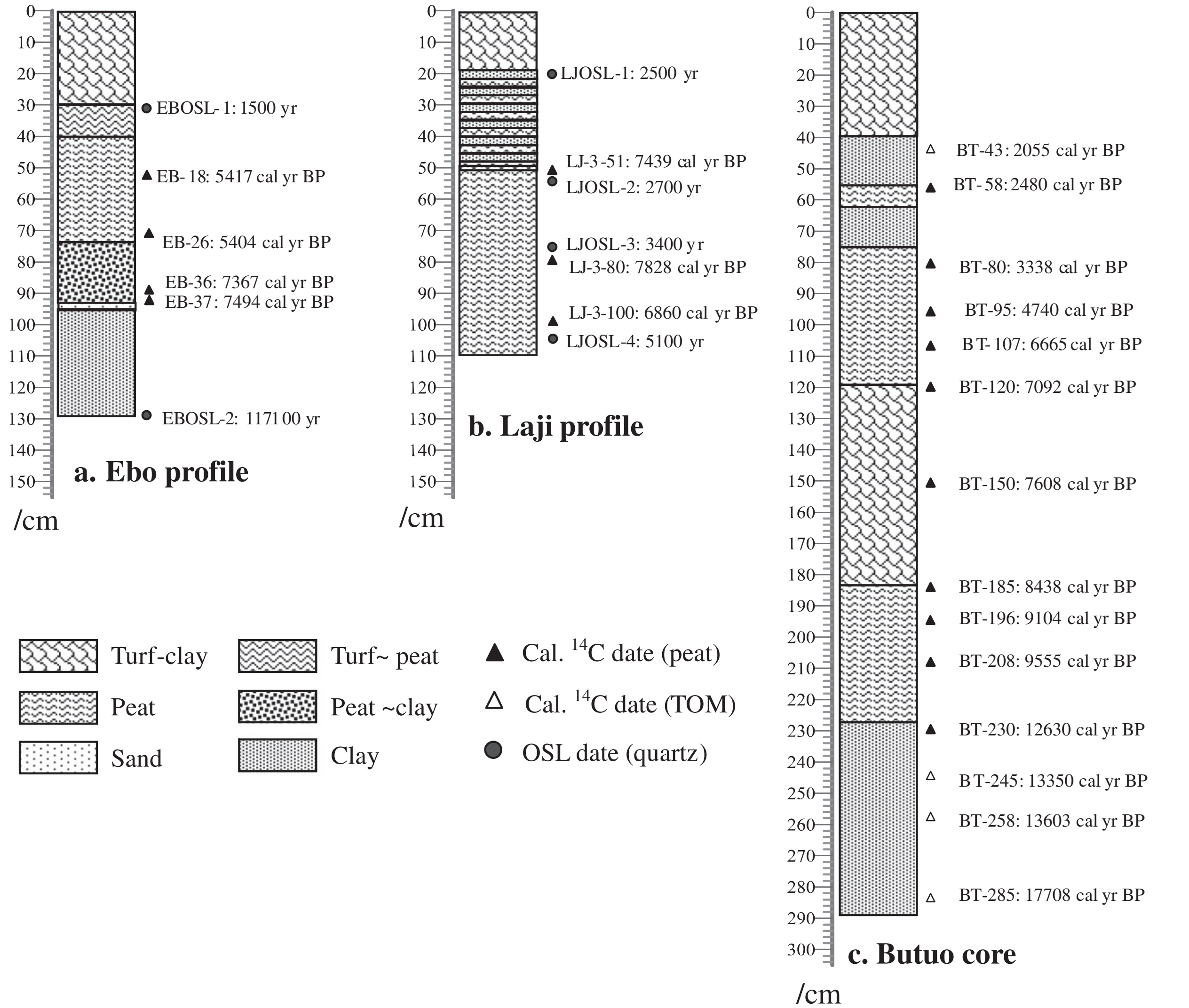

Figure 2. Lithology of the Ebo and Laji profiles and the Butuo core. The calibrated ("Cal.") median ^{14}C ages and the OSL ages are also shown (see Table 1 for details).

Background and methods

Background

The Ebo and Laji profiles are located in relatively low-lying lands surrounded by shallow blanket grass-turf covers within the Qilian and Laji mountains, respectively (Fig. 1). The Ebo profile (37°52′2″N, 101°2′53″E, alt. 3537 m) is located in Qilian county. It has a mean annual temperature of ~1°C and mean annual precipitation of 280–400 mm. The profile is 130 cm in depth, and is underlain by permanently frozen earth. The Laji profile (36°23′19″N, 101°20′28″E, alt. 3843 m) is located in Guide county; mean annual precipitation and mean annual temperature of Guide county is 251–559 mm and ~7.2°C, respectively. The profile is 110 cm in depth, and the underlying material is a mixture of frozen clay and coarse gravels.

The Butuo profile (27°38′34.752″N, 102°55′53.122″E, alt. 2697 m) is located in Le'an village, Butuo county, southwest Sichuan province. Its mean annual temperature is ~10.2°C, and the mean annual precipitation is ~1102 mm. A continuous, undisturbed core of 289 cm depth was extracted from a raised bog within an intermountain basin using a Eijkelkamp core sampler in 2009.

Lithology and sampling

Ebo profile

The uppermost 0–30 cm of the Ebo profile is yellow-brown turf-clay, among which the first 10 cm is strongly influenced by modern depasture and/or human activities, with high clay content. At 31–40 cm is a transition layer of turf and peat, while 41–74 cm is peat and 75–93 cm is peat-clay. At 94–96 cm is a brown sand–clay transition layer, and 97 cm to the end is black-gray clay. Two horizontal undisturbed tube samples were collected at 30 cm and 130 cm for Optically Stimulated Luminescence (OSL) dating, and peat samples at 53–54 cm, 69–70 cm, 89–90 cm, and 91–92 cm were selected for AMS ^{14}C dating (Table 1; Fig. 2).

Laji profile

The surface 0–18 cm of the Laji profile is black turf, with high clay content. At 19–50 cm are alternating layers of yellow sand/clay and black peat. At 51–100 cm is black peat with clay and sparse coarse gravels. OSL samples were collected at 20, 52, 75, and 105 cm; ^{14}C dating samples of peat were collected at 51, 80, and 100 cm (Table 1; Fig. 2).

Butuo peat core

The uppermost 0–26 cm of the Butuo peat core is disturbed turf-clay, and 27–41 cm is yellowish turf. The turf layer is underlain by a gray to gray-brown clayish lake sediment section from 41 to 76 cm, interrupted at 56–63 cm by a thin peat layer. Below this clay section is a relatively thick peat section from 77 to 230 cm. At 77–120 cm is black to gray-black peat, and at 121–185 cm is gray-yellow turf, which is much looser compared with the upper and lower layers. At 186–205 cm and 206–230 cm are gray-black peat and gray peat. From 231 cm to the bottom of this core is lake sediment; 231–255 cm is fine gray clay, while 256–289 cm is clay mixed up with coarse sand. The core was sampled at 1-cm intervals except for the uppermost 26 cm, which is strongly influenced by modern human activities, including animal pasturages. Fourteen samples were selected for ^{14}C measurements, among which four (Butuo-43, -245, -258, and -285) are clay sediments, while the others are peat (Table 1; Fig. 2).

Dating

^{14}C dating materials from clay samples are total organic matter (TOM), while those from peat samples are cellulose (pure organic matter extracted from peat deposits according to Xu et al., 2006). The ^{14}C dating was carried out at both IEECAS and the Beta Analytic Radiocarbon Dating Laboratory (Table 1).

OSL ages were measured using 4–11 μm quartz by a Multiple Aliquot Regenerative-dose method (Lu et al., 2007) at the Institute of Earth Environment, Chinese Academy of Sciences (IEECAS).

Proxy indices analyses of Butuo core

Samples from the Butuo peat core were digested in sodium hydroxide (0.1 M NaOH) to extract humic acid, and the absorbances at 400 nm wavelength were measured on an UV spectrophotometer (SHIMADZU UV2550) to indicate the degree of humification (Blackford and Chambers, 1993; Wang et al., 2010). The remains were neutralized by diluted HCL and rinsed repeatedly in pure water, then dried and ground to powder. Total organic carbon content (TOC) and total nitrogen were measured on an elemental analyzer (vario EL III), with measurement errors less than 0.2%. The loss of organic matter on ignition was measured at 550°C (LOI_{550}) for 2 h, using the raw samples.

Climatic significance of proxy indices

Both LOI_{550} and TOC are widely used to indicate the organic matter content of the sediment. Higher organic matter contents indicate higher peat purity, and thus imply higher plant cover in the catchment, and higher biomass and accumulation rate in the peatlands. We therefore used LOI_{550} and TOC to indicate peatland development in this work.

Vascular plants are rich in fibers but low in proteins, and therefore have high atomic C/N ratios (generally >20) (Meyer, 1997; Xu et al., 2007). On the other hand, the algae/plankton contains less fiber but much more proteins, hence have low atomic C/N ratios (5–12, but generally <10) (Meyers, 1997; Xu et al., 2007). As a result, the C/N ratios are widely used to indicate the relative contribution of terrestrial plants of the catchments and algae/plankton within lakes. Higher C/N ratio indicates higher fraction of terrestrial contribution but lower fraction of algae/plankton, and vice versa.

Humification is generally used to indicate the degree of organic matter degradation (Wang et al., 2010). Our work showed that the humification is highly and positively correlated with LOI_{550} along the ETP transect (Sheng et al., 2013), suggesting that higher humification relates to higher organic matter contents and thus high biomass. However, higher organic matter is not necessarily related to higher humification if the degradation is not concurrently strong.

Results

Chronologies

The median calibrated ^{14}C ages for the Ebo profile are generally between 7500 and 5400 cal yr BP. The OSL ages at 30 and 130 cm are 1.5 ± 0.1 ka and 117.1 ± 5 ka, respectively. These differ from the corresponding ^{14}C ages at a similar depth (Fig. 2a; Table 1). The difference between the OSL age of the upper peat profile and that of the basal permafrost is >100,000 yr, indicating a significant unconformity.

The median ^{14}C ages at 51, 80, and 100 cm of the Laji profile are 7439, 7828, and 6860 cal yr BP, respectively (Table 1; Fig. 2b). The OSL ages range between 5.1 and 2.5 ka, much younger than the corresponding ^{14}C ages. The ^{14}C ages show an inversion with depth; while the OSL ages show a reasonable deposition sequence from young to old as depth increases (Fig. 2b).

The ^{14}C ages of the Butuo core ranged between 17,700 and 2000 cal yr BP (Fig. 2c). Ages from the upper 76 cm are <3100 cal yr BP. The underlying peat section (76–205 cm) is dated to between 3100 and

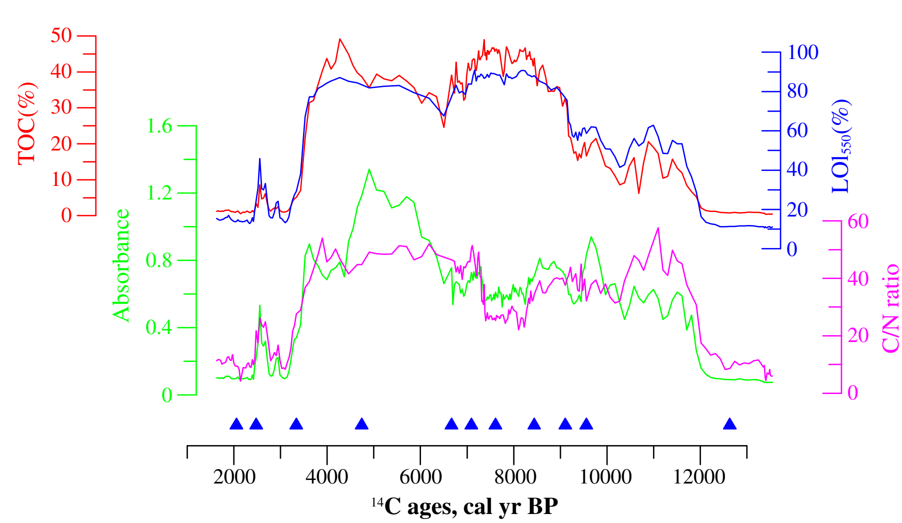

Figure 3. Proxy indices of Butuo peat core (see text for details). Blue triangles show the control points from the ^{14}C dating (median calibrated ages).

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9500 cal yr BP. Ages of the lower peat-clay section (206–230 cm) ranged from 9500 to 11,800 cal yr BP. The lower gray lake sediment section from 231 to 258 cm was dated to between 11,800 and 13,600 cal yr BP. Near the bottom of this core (285 cm) is dated at 17,708 cal yr BP.

LOI, TOC, C/N and humification

LOI_{550} varies between 4.5% and 90.9% in the Butuo peat core (Fig. 3). During 9000–3600 cal yr BP, LOI_{550} was generally high (~70–90%), indicating higher organic matter content, consistent with the peat lithology. During 11,800–9000 cal yr BP LOI_{550} was around 50–60%, consistent with the peat-clay lithology. Before 11,800 cal yr BP and after 3600 cal yr BP, the LOI_{550} was low (generally lower than 20%), consistent with the clay lithology. The high LOI_{550} between 9000 and 3600 cal yr BP can be divided into two parts: 9000–6500 cal yr BP and 6500–3600 cal yr BP. The TOC of Butuo core varies between ~0.36 and 49.18%. TOC and LOI_{550} show very similar patterns for the whole core ($r^2 = 0.95$).

C/N ratios of organic matter in the Butuo core vary between 3.89 and 57.61 (Fig. 3). The C/N ratios during 11,800–3600 cal yr BP are much higher than those of the previous period and of the latter part (3600–2000 cal yr BP). C/N ratios increased significantly after the end of Younger Dryas, and reached a maximum around 11,000 cal yr BP. Variations in C/N ratios are generally similar to those in TOC ($r^2 = 0.5$). However, there are also obvious differences between these two indices. For example, during 11,000–7500 cal BP the TOC shows an increasing trend while C/N ratio shows a decreasing trend, which is possibly due to the climatic changes and the related peatland development (see below).

The degree of humification was much higher during 11,800–3600 cal yr BP (Fig. 3) than subsequently or previously, and was broadly similar to the trends of TOC and LOI_{550}. However, obvious differences exist between humification and TOC and LOI_{550}. For example, from 9000 to 6500 cal yr BP humification was ~constant whereas TOC and LOI_{550} increased remarkably. Humification reached its maximum during the interval of 6500–3600 cal yr BP, while the peak values of TOC and LOI_{550} occurred between 9000 and 6500 cal yr BP.

Discussion

Peatland initiation along the ETP transect

The ^{14}C ages of peat cellulose indicate the time when the organic matter was synthesized, whereas the OSL ages dated by measurements of quartz luminescence signals indicate the latest time when the samples were buried (Lu et al., 2007; Madsen and Murray, 2009), but not necessarily the original formation time of the organic matter. If the peat deposits were never disturbed, then the ^{14}C and OSL ages should be the same, within experimental error. However, if the peat samples were reworked, then the ^{14}C ages would be older than the OSL ages. Accordingly, a combination of ^{14}C and OSL ages can provide more information on the history of peatlands initiation and subsequent erosion than either can provide by itself.

The ^{14}C ages of the Ebo profile range between 7500 and 5400 cal yr BP, falling within the traditional Holocene thermal optimum (e.g. 8500–3000 ^{14}C yr BP [9400–3100 cal yr BP]; Shi et al., 1992; Zhao et al., 2009). The >100 ka OSL age of the underlying frozen earth suggests either that there was no peat accumulation during the time of the last glacial stades, and/or that the peatlands formed during the interstades were completely eroded. The OSL age at 30 cm is ~1.5 ka, much younger than the corresponding interpolated ^{14}C age, suggesting the sediments were not originally formed in situ, but represent re-deposition after erosion and removal.

The ^{14}C ages of Laji profile range between 8000 and 6500 cal yr BP, also suggesting that the peat growth was initiated during the Holocene optimum. The reversal of the ^{14}C ages, but not of the OSL ages (<5100 yr), as well as the remarkable differences between ^{14}C ages and the corresponding OSL ages (Fig. 2b; Table 1), suggest that at ~5100 yr ago the peatland stopped net accumulation, and the peat was then eroded, transported, and re-deposited elsewhere.

The ^{14}C ages of the Butuo peat show that the initial age of Butuo peatland is about 11,800 cal yr BP, soon after the end of Younger Dryas and coinciding with the initiation of Hongyuan peatlands (Hong et al., 2003; Xu et al., 2006; Zhou et al., 2010; Zhao et al., 2011). The favorable period for peatland expansion inferred from the TOC and LOI_{550} proxy indices is 9000–4000 cal yr BP. During 11,800–9000 cal yr BP, peatland formation rates were probably low, as deduced from the low TOC and LOI_{550} values and the high clay content. After ~3600 cal yr BP, peat accumulation ended, and was replaced by clay sedimentation. This cessation in peatland development is generally synchronous with the decline of peat accumulation rate at the Zoige plateau, northwest Sichuan province (see below).

Response of Butuo peatland to climatic changes

As shown in Fig. 3, the high peat biomass at Butuo inferred from the TOC and LOI_{550} data occurred 9000–4000 cal yr BP, within the Holocene climatic favorable period near/around this region as inferred from the stalagmite $\delta^{18}O$ in Dongge cave (Dykoski et al., 2005) and from the $\delta^{13}C$ of the Hongyuan peat profile (Hong et al., 2003). Zhao et al. (2011) compared their pollen data at Zoige plateau with previous work along the ETP transect and showed a favorable plant growth period during about 10,500–3600 cal yr BP. These suggest that the plant biomass was high during the Holocene favorable period, broadly consistent with the Holocene thermal optimum period (Shi et al., 1992; An et al., 2000).

The ^{14}C ages and proxy indices suggest that the most favorable period for Butuo peat growth was 9000–6500 cal yr BP. The fast accumulation of organic matter then can be partly attributed to the increases in photosynthesis due to higher precipitation. The C/N ratios during this interval decreased slightly (Fig. 3), implying a more water-logged environment that favors a higher fraction of biomass within the peatland. Generally, a high water table in the peatlands creates anoxic conditions, inhibits aerobic bacteria activities, and therefore lowers the rate at which organic material is decomposed (Clymo, 1984; Ise et al., 2008). As a result, the raised water table in Butuo peatland during 9000–6500 cal yr BP would also contribute to an increased biomass accumulation rate by reducing the organic matter decomposition rate. This lowered decomposition can be corroborated by the relatively steady variations in humification during this period of time (Fig. 3).

An obvious phase lag between the growth optimum and climatic optimum at the Butuo peatland can be seen from the comparisons in Fig. 4. For example, precipitation increased rapidly from about 11,600 to 9000 cal yr BP (Hong et al., 2003; Dykoski et al., 2005), following the end of the Younger Dryas. However, the Butuo TOC and LOI_{550} indices increase slowly and do not reach their Holocene optimum values until ~8000 cal yr BP. The pollen data (e.g. Zhou et al. 2010) also imply that during the early Holocene the vegetation cover on the eastern margin of the Tibetan Plateau was systematically lower than that of the following interval (Fig. 4). This response lag between ecosystem and climate may be attributed to the development of soils and vegetation cover. During the early Holocene, the soils on the eastern margin of the Tibetan Plateau were generally thin due to the strong physical weathering in the glacial times; this resulted in low water tables and low biomass but relatively fast decomposition of organic matter within the catchments. As soils became thicker, the water-holding capacity increased, the biomass increased and decomposition relatively decreased, which eventually led to maximum peatland development. Therefore, it is reasonable that a response lag existed between the climatic optimum and the maximum peatland development. The differences between "TOC, LOI_{550}" and "absorbance, C/N ratios" of the Butuo core during

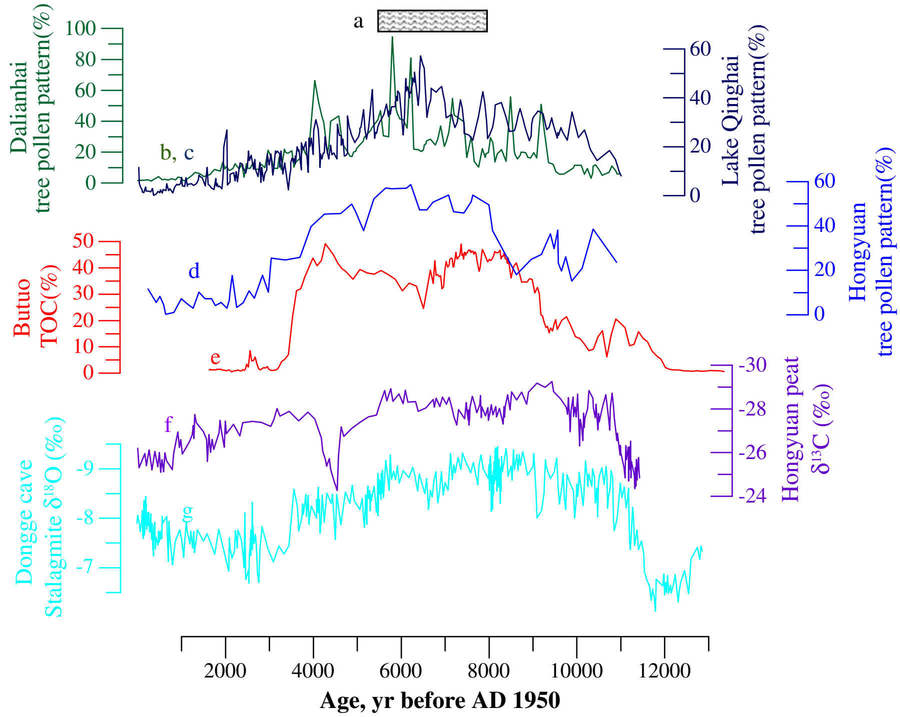

Figure 4. Comparison of the peatland development along the ETP transect. (a) The filled rectangle shows the general period (8000–5400 cal yr BP) of peatland development of Ebo and Laji profiles. (b), (c), and (d) Tree-pollen histories for Lake Dalianhai, Lake Qinghai, and Hongyuan peatland, respectively, redrawn from Zhao et al. (2011). (e) TOC for the Butuo peat core. (f) $\delta^{13}C$ history from the Hongyuan peat core (Hong et al., 2003). (g) $\delta^{18}O$ history from a stalagmite in Dongge cave (Dykoski et al., 2005). See locations of each site in Fig. 1.

the early Holocene (Fig. 3) strongly supports such an "ecology ~ climate" response lag. The phase differences between the tree-pollen pattern from Hongyuan peat deposits (Fig. 4d) and precipitation inferred from Hongyuan peat $\delta^{13}C$ (Fig. 4f) also support such an inference. Zhao et al. (2009) compared the vegetation changes over a wider geographic extension and they also found that gradual vegetation changes in the early Holocene lagged about 1000 yr behind the summer monsoon maximum.

At about 6500 cal yr BP the Butuo peat formation rate decreased slightly, and then increased again until 4200 cal yr BP. At about 4000 cal yr BP the peatland accumulation declined, and stopped at about 3600 cal yr BP. At that time the Butuo indices and the lithology show abrupt changes that are generally consistent with the abrupt vegetation shift from Pinus (decreased to 15–0%) to Artemisia (up to 80%) at ~4.2 ka as suggested by Zhao et al. (2009). The results of our previous work (Hong et al., 2003) and the pollen assemblages by Zhou et al. (2010) and Zhao et al. (2011) also suggested both an obvious decrease in the net peat accumulation rate and an obvious decline in vegetation cover, respectively, on the Zoige plateau during the late Holocene.

Phadtare (2000) examined the pollen composition in a peat profile from northern India, and also found that between 4000 and 3500 cal yr BP, the abundance of conifers sharply decreased, with the greatest increase in evergreen oak, which suggests a sharp decrease in summer monsoon intensity during 4000–3500 cal yr BP. This decreased monsoon precipitation during 4000–3500 cal yr BP also correlates with proxy climatic data from other localities in the Indian subcontinent and western Tibet (Phadtare, 2000), suggesting that this dry episode was a widespread event of the Holocene record

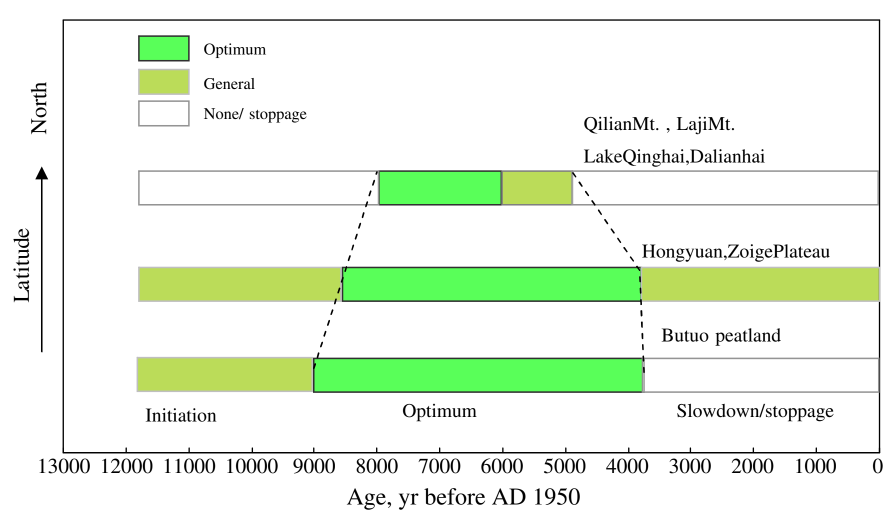

Figure 5. A sketch map of the general peatland development along the ETP transect.

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in south-central Asia. In addition, the striking increase in $\delta^{18}O$ and the sharp decrease in growth rate of a stalagmite in Dongge cave suggested a strong decrease in monsoon precipitation around this same period of time (Dykoski et al., 2005), and the growth termination around 4000 years ago of a stalagmite in the central Tibet plateau also suggested a decrease in summer monsoon intensity then (Cai et al., 2012). We speculate that this monsoon failure (Fig. 3) led to decreased biomass and increased decomposition of organic matter, which eventually resulted in a decline or cessation in peat accumulation.

Temporal and spatial differences of peatland development along the ETP transect

Our work showed that the basal ^{14}C ages of the peatlands of the northeastern ETP are younger than 8000 cal yr BP and peat ages are mainly concentrated between 8000 and 6500 cal yr BP (Fig. 2; Table 1). This suggests that there was little net peat accumulation before 8000 cal yr BP and that the peat growth optimum of the northern ETP was between 8000 and 6500 cal yr BP. The pollen record at Lake Qinghai (Shen et al., 2005) and Lake Dalianhai (Zhao et al., 2011) within this region, also show that the Holocene optimum plant growth period generally occurred between 8 and 6 ka (Fig. 4), consistent with the development pattern of the peatlands.

Peatland initiation in both the middle ETP and southern ETP occurred about 11,800 cal yr BP (Fig. 3), soon after the end of Younger Dryas. This is much earlier than in the northern ETP (Fig. 5). The optimum peat growth time of the northern ETP is also about 1000 yr later than in the middle and southern ETP (9000–6500 cal yr BP), and the duration of the period of optimum growth is much shorter than in the middle ETP and southern ETP (Fig. 5). The later initiation and the lagged optimum of peatland growth in the north are likely due to the cooler and dryer climatic background over the north ETP.

Conversely, the timing of peatland decline in the northern ETP is earlier than in the middle and southern ETP (Fig. 5). Few ^{14}C ages from the Laji and Ebo profiles are <6500 cal yr BP, suggesting a slowdown or cessation of peatland growth then. Pollen records from the northeastern Tibet–Qinghai Plateau also reveal a widespread decline in forest cover after 6 ka. For example, Zhao et al. (2009) showed obvious "*abrupt palynological changes based on a squared-chord distance of pollen assemblages*" at 6–5 ka. The decline of the northern ETP peatlands is possibly because the Asian summer monsoon intensity decreased to a limiting level which cannot support extensive biomass. We therefore suspect that the present sparse peatlands over the northern ETP occupy only a small fraction of the peat landscapes that were present during the Holocene optimum. In contrast, the southern ETP can still support a continuous peatland growth after 6500 cal yr BP. For example, the Butuo peatlands in the southern ETP had another period of optimum growth during 6500–4200 cal yr BP.

As mentioned above, the Butuo peatland stopped net accumulation after about 4000 cal yr BP, while on the Zoige plateau the Hongyuan peatland continued growing (also shown in Fig. 5). We attribute this to the different regional climatic conditions. Evaporation over the southern ETP is stronger than over the Zoige plateau due to the higher air temperature, which leads to much lower soil moisture at the southern ETP than at the middle ETP (like the Zoige plateau). As a result, the peatlands declined/stopped net accumulation in the southern ETP after ~4000 cal yr BP while continuing at the Zoige plateau. This can also be supported by the modern meteorological records. Although precipitation is relatively high over the southern ETP, the strong evaporation there leads to severe aridity (e.g., in northwestern Yunnan province), which strongly impacts the local/regional agriculture and ecosystem. In contrast, farther north the evaporation is lower: therefore, on the Zoige plateau, although the precipitation is relatively low, the soil moisture and thus the vegetation cover are high.

Possible carbon source/sink effect of peatlands along the ETP transect

During the Holocene optimum, the net accumulation suggests that the carbon effect of peatlands along the ETP transect should be a sink. Higher precipitation and temperature should result in a higher rate of net carbon sequestration. As the Asian summer monsoon decreases, some of the peatlands initiated during the Holocene optimum, especially over the northern ETP, ceased to grow and even disappeared due to the decreased biomass and enhanced organic matter degradation. Thus, parts of the carbon captured previously have been returned or are now returning to the atmospheric CO_2 pool. These peatlands should be regarded as net C sources. Living peatlands, like the Zoige wetlands, are still carbon sinks although their accumulation rates are reduced. However, if current aridity continues or increases, these peatlands may deteriorate and become carbon sources. Our most recent work (Xu et al., 2012) showed that the monsoon intensity decreased obviously during the past 100 to 200 years over the Indian summer monsoon region, and suggested that the precipitation would continue to decrease if global warming continues. Given this scenario, the peatlands along the ETP transect would be expected to release more of the previously sequestered CO_2 back to the atmosphere.

Summary

This study addressed the development of Holocene peatlands along the eastern margin of the Tibetan plateau. The general development trend was found to correlate with that of the summer monsoon precipitation. This suggests that the peatland development is mainly dominated by precipitation, which is different with the developments of some other peatlands outside the Asian summer monsoon regions, e.g. the high-latitude northern hemisphere peatlands. The Holocene peat-growth optimum along the whole transect is generally concentrated within the period of about 9000–6500 cal yr BP. This growth optimum lagged the Holocene climatic optimum. Furthermore, the initiation of peat formation over the northern ETP is much later than those over the middle to southern ETP, and the slowdown or cessation in the north is earlier. We attribute these asynchronies between ecosystem and climates to the variations in summer monsoon intensity, and to the local hydrology and geographical characteristics.

Parts of the previously formed peatlands have disappeared and some have ceased net accumulation or are now being eroded. These peatlands returned and/or are now returning the previously captured carbon to the atmosphere and therefore act as net carbon sources. Some peatlands still have positive net accumulation at present, but their accumulation rates are now reduced, concomitant with the long-term decreasing Asian summer monsoon intensity. These living peatlands still act as net carbon sinks. However, if the aridity continues, as is possible with continued global warming, even more of the carbon previously stored in the peatlands of the ETP would be released to the atmosphere.

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